JP4.5 CHARACTERIZATION AND VALIDATION OF THE HEAT STORAGE VARIABILITY FROM TOPEX/POSEIDON AT FOUR OCEANOGRAPHIC SITES

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1. INTRODUCTION

Estimates of the oceanic heat storage (HS) are important to improve our understanding of air-sea interactions and its long-term contribution to climate change. HS is traditionally estimated from *in situ* measurements of temperature at specific locations (Yan et al. 1995) or on climatological temperature data sets (Moisan and Niiler 1997). Most recently, HS has been estimated using altimetry data (White and Tai 1995; Wang and Koblinsky 1997; Chambers et al. 1997). The satellite altimeter measures the total sea level anomaly, with contributions from many different phenomena. The objective of this study is to estimate the HS using TOPEX/Poseidon (T/P) data, identifying the contributions from the thermal and dynamical processes involved in the HS mechanisms. HS and its rate (HSR) are estimated from *in situ* data and compared to the results based on remotely sensed data.

2. METHOD

The HS of an observed temperature profile is:

$$HS = \rho C_p \int_{-h}^{0} T(z)dz, \tag{1}$$

where ρ is the density of seawater, C_p is the specific heat at constant pressure, $T(z)$ is the temperature profile, and h is the depth to which the temperature is integrated. HS is expressed in units of $J.m^{-2}$. The HSR estimated using the altimeter data is determined by differentiating HS$'$ in time.

The heat storage anomaly (HS$'$) is estimated from the filtered height anomaly (η) according to a linear relation (Chambers, Tapley, and Stewart 1997):

$$HS' = \frac{\rho C_p}{\alpha}(\eta + \eta_h), \tag{2}$$

where α is the thermal expansion coefficient. In this study, the effects of the haline contraction (η_h) on the sea height anomaly are investigated.

Mean values of ρ and C_p, averaged from the surface to a depth h, are calculated from climatological maps of the World Ocean Atlas 1994 (WOA94) (Levitus and Boyer 1994) for a $1° \times 1°$ grid. α is considered constant at each grid point and it is estimated by vertically averaging between the surface and the depth h the climatological α profile weighted by layer thickness and by temperature anomaly. η_h is estimated by vertically integrating the product of the climatological haline contraction coefficient, β, and the salinity anomaly (in relation to the annual mean) profiles.

The T/P sea surface height anomaly is decomposed using 2D finite impulse response filtering (Polito and Cornillon 1997). This method separates the zonal-temporal signal $\eta(x,t)$ into additive components:

$$\eta = \eta_t + \eta_w + \eta_r, \tag{3}$$

η_t is the basin-wide non-propagating variability, mostly due to seasonal heating and cooling and advection by the broad oceanic currents. η_w is the large to meso-scale westward propagating signal composed mainly of first-mode baroclinic Rossby waves, with periods of 24, 12, 6, 3 and 1.5 months. η_r includes a variety of signals among them equatorial Kelvin waves and meso-scale eddy variability. The small-scale, non-propagating signals are filtered out.

3. DATA

Four sites were selected (Table 1): TAO array in the equatorial Pacific, the hydrographic sections from the California Cooperative Oceanic Fisheries Investigations (CalCOFI) cruises at the California coast; the hydrostation ALOHA from the Hawaii Ocean Time series Program (HOT) in Hawaii; and the hydrographic time series in Bermuda (Hydrostation "S") in the western Atlantic. The general procedure consists of linearly interpolating individual temperature and salinity profiles in the vertical.

Table 1: Data source, number of stations used (N), average sampling rate and their location.

Source	N	Rate (days)	Location
TAO	50	1	$8°S - 9°N$, $140°E - 95°W$
CalCOFI	11	90	$29.5°S - 35°N$, $116°W - 124°W$
HOT	1	40	$22°N, 158°W$
Hyd. "S"	1	15	$32.2°N, 64.5°W$

The T/P data are the JPL/PODAAC WOCE global from 10/1992 to 12/1998 interpolated to a 1°× 1°× 10 days grid by a bicubic interpolator.

Tidal aliasing is a potential problem caused by the temporal under-sampling of the remainder of the tidal signal left in the T/P record after the model tides are removed. (Schlax and Chelton 1994) calculated the aliased periods and wavelengths for the major tidal constituents. The majority of the aliasing problems affect η_1. Analogously, Rossby waves can be aliased whenever the wavelength is similar or smaller than twice the track separation. To avoid this problem the filter for each frequency was limited to a latitudinal band specific to that frequency.

4. RESULTS

An example of the composition of the T/P altimeter signal for the TAO region is shown in Figure 1 for 4.5°N. Slanted patterns indicate propagation. The most important point is that the sum of the filtered fields η_s is a good approximation of the original field η_o, accounting for over 90% of the total variance. The amount of variance σ explained by η_t is larger than the sum of the wave signals and it grows towards the Equator.

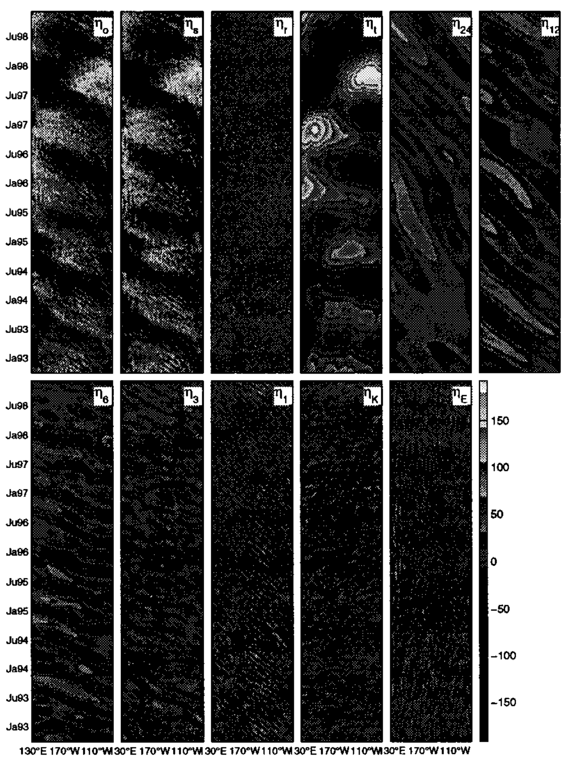

Figure 1: Sea surface height anomalies components (in mm) from T/P as a function of longitude and time at 4.5°N in the Pacific (TAO). The color scale is the same for all plates.

Most westward propagating signals ($\eta_{24,12,6,3}$) can, in principle, be identified as long equatorial Rossby waves; η_1 corresponds to tropical instability waves. The eastward propagating signal (η_K) has both phase speed and period that match those of equatorial Kelvin waves.

Estimates at four selected buoys are compared to the T/P estimates, Figure 2. The HSR (right panel on Figure 2) is calculated with a time differential of 30 days. This time interval helps to counteract the deterioration of the signal to noise ratio due to differentiation. The mean HSR rms difference between both estimates is $149\,W m^{-2}$ and the correlation is 70%. The inclusion of a climatological salinity effect did not change the results significantly. This suggests that for the TAO region the seasonal fluctuations in the climatological salinity are not the dominant salinity signal.

The HS′ error based on the error of the temperature measurements in the TAO data alone is $50 \times 10^7\,J m^{-2}$ (Chambers et al. 1997). The mean rms difference in this study is $54 \times 10^7\,J m^{-2}$ and a mean correlation of 88%.

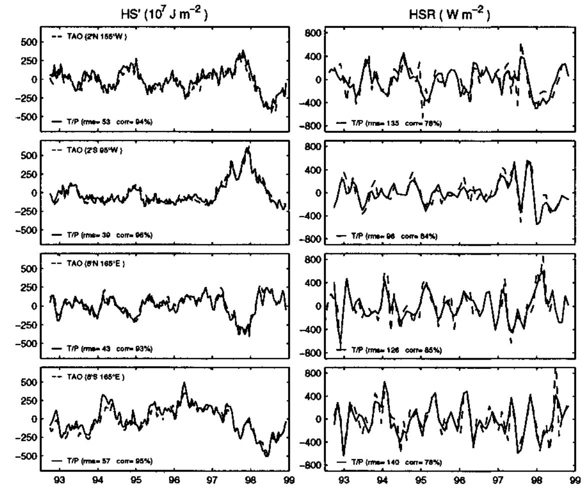

Figure 2: Comparison of the heat storage anomaly (left) and heat storage rate between TAO (dashed) and T/P (solid) at selected buoys. The rms difference and correlation coefficients are shown in the lower left.

For the Hydrostation "S" region the thermal signal and the sum of the propagating signals explain approximately the same amount of variance. This is an important result which underlines the importance of Rossby waves for local estimates of the heat budget.

The dominant signal in the heat storage spectrum apparently shifts from semi-annual before 1995 to annual after. Results are in better agreement after 1995: lower rms differences and higher correlation ($53 \times 10^7\,J m^{-2}$ and 86%) are obtained for the 1995–1997 period compared to 1993–1995 ($69 \times 10^7\,J m^{-2}$ and 72%) (Figure 3). The location of

Hydrostation "S" coincides with a T/P cross-over latitude and T/P cannot properly resolve the semi-annual signal which results in spatial aliasing.

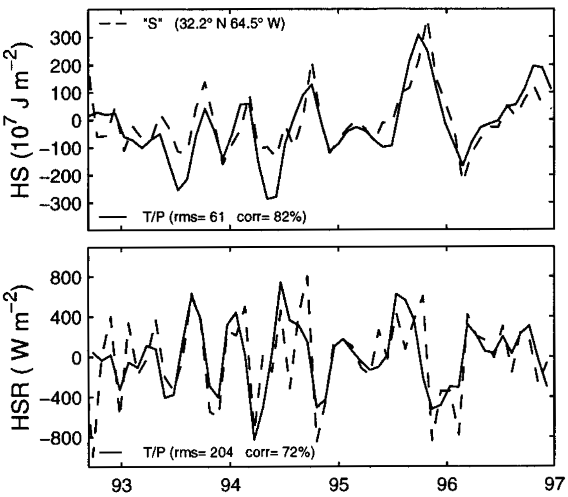

Figure 3: Comparison of the heat storage anomaly (top) and heat storage rate (bottom) between Hydrostation "S" (dashed) and T/P (solid).

The rms difference between the two HS′ estimates is 61 $\times 10^7\, J\, m^{-2}$ and the correlation is 82%. A comparison of the two HSRs shows a rms difference of 204 $W\, m^{-2}$ and a correlation coefficient of 72%. The fractional variance of η_t is a factor of 2 smaller than that of Hydrostation "S" and a factor of 5 smaller than

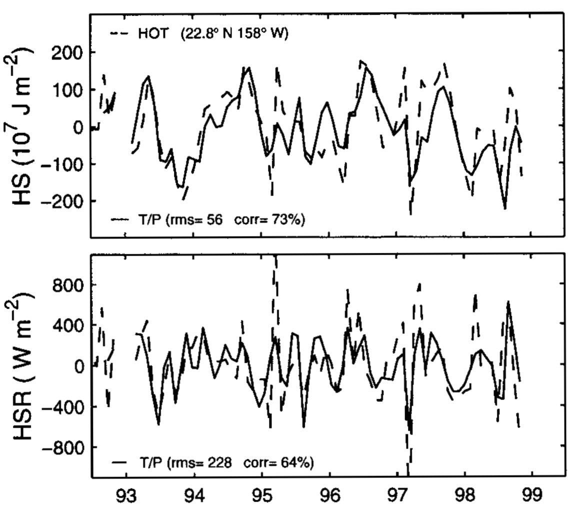

Figure 4: Comparison of the heat storage anomaly (top) and heat storage rate (bottom) between HOT and T/P, similar to Figure 3.

that of TAO. The total Rossby wave signal explains over 4 times more variance than η_t, with η_6 being the strongest.

The HS′ and HSR from HOT and T/P show reasonable correlations and rms differences. The major source of uncertainty of the HS′ estimates derives from the sampling rate of the *in situ* data. The average rate of one sample every 40 days is marginally able to sample η_3, responsible for 19% of the variance (Figure 4). The rms difference between the hydrographic and altimetric estimates of HS′ is 56 $\times 10^7\, J\, m^{-2}$ and the correlation coefficient is 73%. Similarly for the HSR the rms difference is 228 $W\, m^{-2}$ and the correlation is 64%.

The non-propagating signal (η_t) at CalCOFI has a fractional variance similar to that estimated for the Hydrostation "S". The total westward propagating signal explains approximately as much of the variance as the non-propagating ($\sim$ 30%). CalCOFI is the location with, on average, the smallest correlations and largest rms differences of all *in situ* data sources used in this study. Several factors contribute to the discrepancies. The temporal resolution of one sample every 90 days can barely resolve η_6.

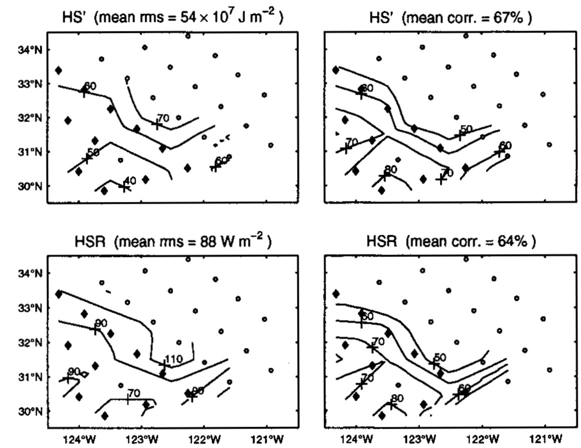

Figure 5: Rms differences (left) and correlation coefficients between CalCOFI and T/P estimates of heat storage anomaly (top) in $10^7 Jm^{-2}$ and heat storage rate (bottom) in Wm^{-2}. Diamonds (circles) mark the location of the used (discarded) stations.

Satellite measurements of the sea surface height degrade near the coast due, to a large extent, to local tides that are inadequately modeled in the T/P data and spread westward by the filter. Thus, correlations decrease and rms differences increase towards the coast. A strong gradient in both rms and correlation is located in the approximate SE-NW diagonal of Figure 5. Therefore, only eleven stations west of this gradient were considered. Station 90/110 located at 30.75°N, 123.33°W gave results which were de-correlated with all stations in its vicinity and was excluded from the analysis. The CalCOFI stations have lower temporal resolution compared to the other sites, with one sample every 90 days. The T/P time series was interpolated to match this resolution. The HS′ mean rms difference and correlation for the region are $54\times10^7 Jm^{-2}$ and 67% while for the HSR they are $88\,W\,m^{-2}$ and 64%.

5. CONCLUSIONS

The Rossby wave aliasing problem is to the best of our knowledge an original finding. This problem is more pronounced at high latitudes (higher than ~35°) and affects estimates of amplitude, wavelength, and phase speed of first mode baroclinic Rossby waves.

The TAO region shows a high correlation between the heat storage estimated from *in situ* and T/P time-series. The rms differences are in average $54\times10^7\,J\,m^{-2}$ and the mean correlation is 88%. The heat storage rates have a mean rms difference of $149\,W\,m^{-2}$ and a mean correlation of 70%. The sea surface height anomaly in this region is dominated by the thermosteric non-propagating signal which near the Equator is responsible for approximately 75% of the T/P variability; towards higher latitudes the propagating signals become more important than the non-propagating ones.

The comparison of HS′ estimates from Hydrostation "S" near Bermuda and T/P shows an rms difference of $61\times10^7\,J\,m^{-2}$ and a correlation coefficient of 82%. The rms difference between the HSRs is $204\,W\,m^{-2}$ with a 72% correlation. The location of Hydrostation "S" coincides with a crossover latitude aggravating the Rossby wave aliasing problem. At this location the basin-scale signal accounts for 35% of the variance while Rossby waves are responsible for 29%.

For the HOT station the rms difference is $56\times10^7\,J\,m^{-2}$ and a correlation coefficient of 73% for the HS′ and $228\,W\,m^{-2}$ and 64% for the HSR. However, the main problem in this case is the temporal resolution of 40 days in a region where Rossby waves with a period of 90 days are responsible for 19% of the variance. The spectrum is dominated by planetary waves which account for 62% of the variance in contrast to 13% associated with non-propagating basin-scale signals.

The mid-latitude northeastern Pacific (CalCOFI) data yields an average HS′ rms difference of $54\times10^7\,J\,m^{-2}$ and a correlation coefficient of 67%, while the HSR has an average rms difference of $88\,W\,m^{-2}$ and a correlation of 64% in comparison with the T/P estimates. The correlation is the lowest of all studied sites. This is due in part to the proximity to the coast. In addition, the relatively long sampling period (90 days) and the presence of a crossover latitude within the sampled area detract from the results. On average at this location the planetary waves are predominant, explaining 35% of the variance, half of it in the semiannual band. The basin-scale thermal signal is responsible for 30% of the variance.

This comparison with *in situ* data helps to validate the altimeter as a valuable tool to study climate-relevant variables in the ocean in detail. This study underlines the importance of wide-spread long-term *in situ* measurements for the remote-sensing community as well as the necessity of a continuous monitoring of the sea surface height by satellite altimeters.

6. REFERENCES

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